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ANALYTIC MODEL FOR ZONAL WINDS IN THE TROPICS

II. Variation of the Tropospheric Mean Structure With Season and Differences Between Hemispheres

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ABSTRACT

An equatorial β-plane model for the tropospheric zonal circulation is used to examine the consequences of the seasonal and hemispheric variation both of the tropical rain belt as a zonal mean heat source and of the horizontal eddy momentum fluxes as a zonal mean momentum source. The model calculations show variations of Hadley circulations with hemisphere and season. The winter hemisphere Hadley cell is more intense in July than in January because of the greater mean displacement of the tropical rain belt from the Equator and hence the greater asymmetry between hemispheres. The essential differences between the zonal winds in the two hemispheres during summer are reproduced by differences in the eddy momentum transports and in the mean meridional circulation.

The model indicates how the annual oscillation of temperature in the equatorial lower stratosphere, with lowest temperatures in January, derives from the difference between the upward branches of the July and the January winter hemisphere Hadley cell. The semiannual oscillation in winds and temperatures in the Tropics is largely accounted for by the model in terms of the longitudinally averaged tropical rain belt migrating between summer hemispheres.

1. INTRODUCTION

In discussing the seasonal changes of the tropospheric zonal mean state, we consider a stably stratified ideal gas, subject to the equations governing fluid motions on an equatorial β-plane. The time scales for frictional and radiative damping are short compared to a year, so the structure of the zonal mean winds and temperatures is largely determined by the balance between sources and sinks of heat and momentum. We distinguish here between sources and sinks that can be prescribed from observations and those regarded as dissipative, in other words, frictional drag and radiative damping. The frictional drag in the absence of momentum sources restores the zonal wind to the same velocity as the underlying surface, whereas the radiative damping in the absence of other heat sources restores the temperature at a given pressure level to some constant reference temperature.

It is clear that the zero level of the source distribution must be properly defined at each point if the damping processes are to restore the atmosphere to the above-mentioned basic states. It is also recognized that all the zonal mean sources of heat and momentum with the exception of incoming solar radiation are actually related in some way to the zonal mean variables. Those sources are assumed given in which the relationship to the zonal mean is considerably more complicated than just a relaxation of the zonal mean variables back to a basic state.

It is our purpose here to isolate the role of these relatively complicated sources in determining the zonal mean state in the troposphere. Our model especially emphasizes the importance of the tropical maximum of latent heat release and of the horizontal transports of momentum by large-scale eddies. For a more fundamental understanding, it is necessary, in turn, to know the role of the zonal mean state in structuring these sources. We believe it helpful to study those questions separately before considering the dynamics of the source-coupled system in all its complexity. Here, we seek to describe diagnostically the observed zonal mean circulation in terms of the balances that exist in the equations for the zonal mean variables including observed sources of heat and momentum.

In part I (a companion paper by Dickinson 1971, hereafter referred to as I), we showed how the Hadley circulation forced by the tropical rain belt creates westerly momentum in the tropical upper troposphere. This momentum is transported by eddies to middle latitudes, thus maintaining the maximum westerly jets there. Forced meridonal circulations, in turn, redistribute heat and momentum so that the resulting zonal winds and temperatures remain in thermal wind balance. Our purpose now is to discuss how quantitative differences in these balances can account for the observed differences of the zonal mean state from one season to another and between hemispheres. We are again restricted to latitudes within 45° of the Equator because of the limitations of our model. Our discussion in all cases applies only to the longitudinally averaged component of the circulation.

2. OBSERVED ZONAL MEAN STATE TO BE EXPLAINED

Perhaps the most familiar zonal wind change with season is the equatorward shift and strengthening of the zonal westerlies from summer to winter as seen in the January and July mean geostrophic wind sections of figure 1.[2] Also of note in the lower troposphere is the

[1] The National Center for Atmospheric Research is sponsored by the National Science Foundation.

[2] Figures 1 through 4 were prepared by H. van Loon and R. L. Jenne from material in van Loon and Jenne (1972).

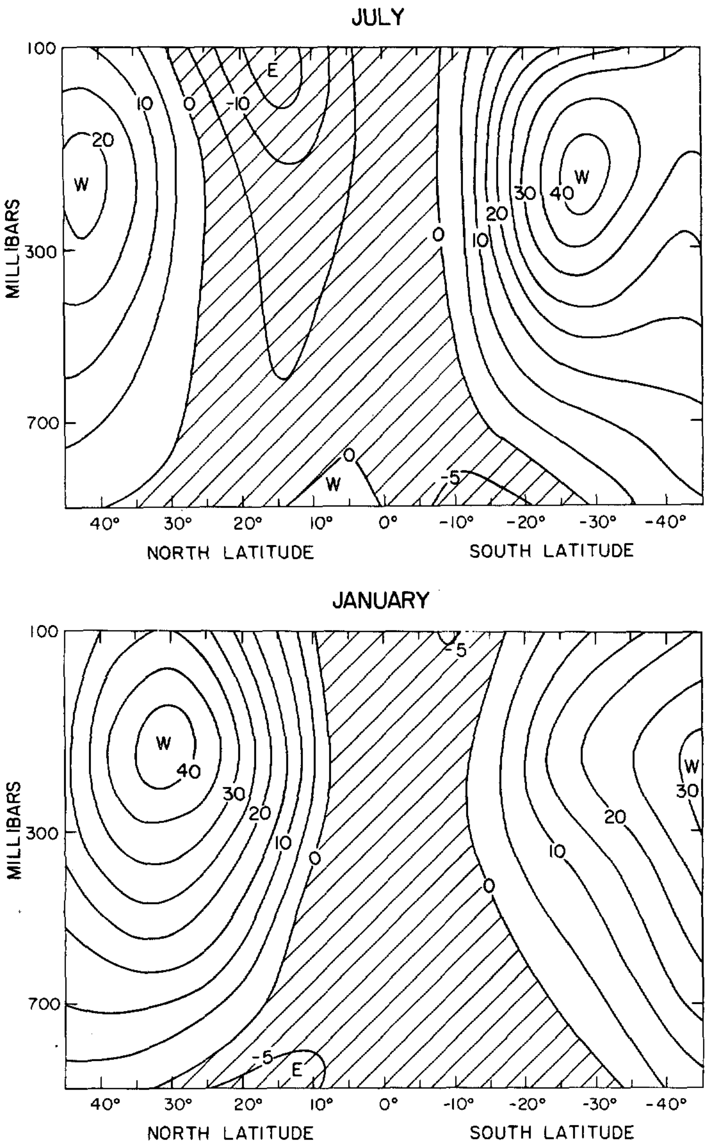


FIGURE 1.—Observed mean geostrophic zonal wind for July and January from the Tropics to middle latitudes; units, m/s.

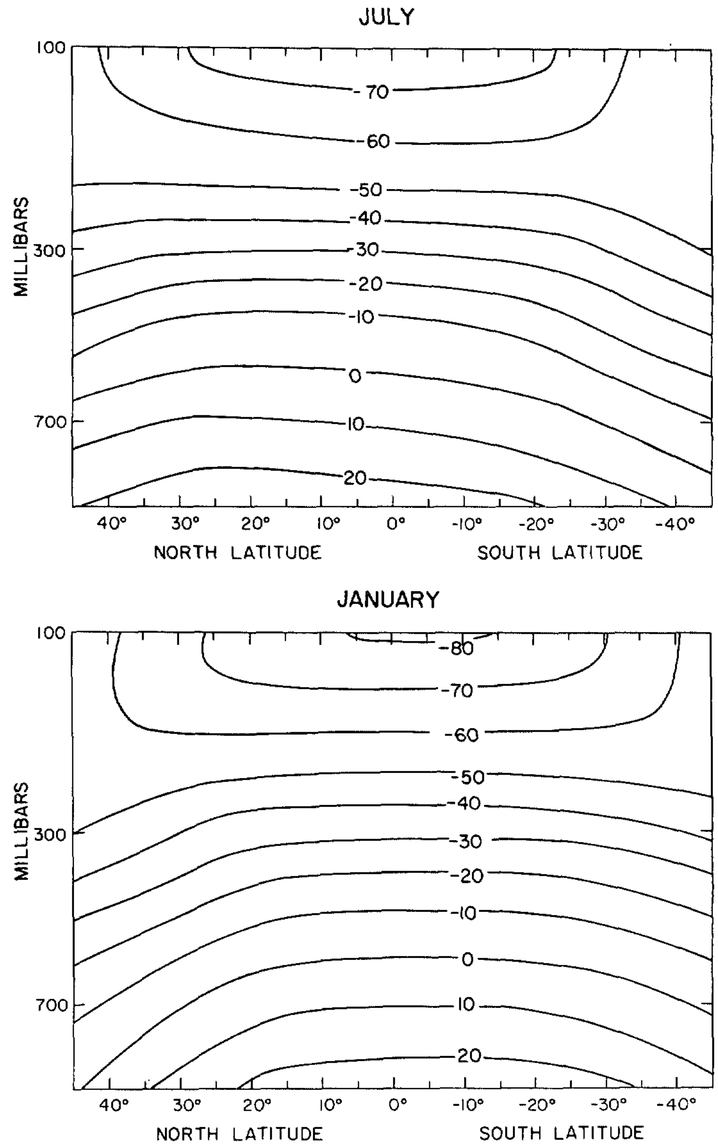


FIGURE 2.—Observed mean temperatures for July and January from the Tropics to middle latitudes; units, °C.

greater strength of the subtropical easterlies in winter compared to that in summer and the presence in the northern summer of equatorial westerlies. The most striking differences between hemispheres in the troposheric zonal winds, as seen in figure 1, are the relative weakness of the westerly jet and the relative strength of the downward extension from the stratosphere of the tropical easterly maximum in the Northern Hemisphere summer compared to the Southern Hemisphere summer.

The observed zonal mean temperature fields, shown in figure 2, indicate a minimum at the tropopause in equatorial latitudes which is colder (4°C lower) in January than in July. This temperature anomaly extends upward into the stratosphere. The bulk of the troposphere at the Equator, on the other hand, is generally warmer (almost 1°C higher) in January than in July. Also of note is the region of relative warmth at latitudes 30° to 40° in the winter hemisphere, lower stratosphere, best seen in figure 2 from the latitude at which the −60°C contour intersects the 100-mb level. This region is warmer in the Southern Hemisphere in July than in the Northern Hemisphere in January (van Loon and Jenne 1969). The reader may also refer to 3-mo mean values of Kidson et al. (1969), their figure 3 for "real" zonal winds, and their tables 1 and 2 for temperatures. All of the abovementioned features are again evident, with the exception of the summer equatorial westerlies in July in the lower troposphere.

The seasonal variations of zonal winds and temperatures are brought out in greater perspective by a harmonic analysis in time. Figure 3 shows the first harmonic in winds and temperatures, and figure 4 shows the second harmonic in these variables. The details of phase variation have been simplified by taking all phases to be

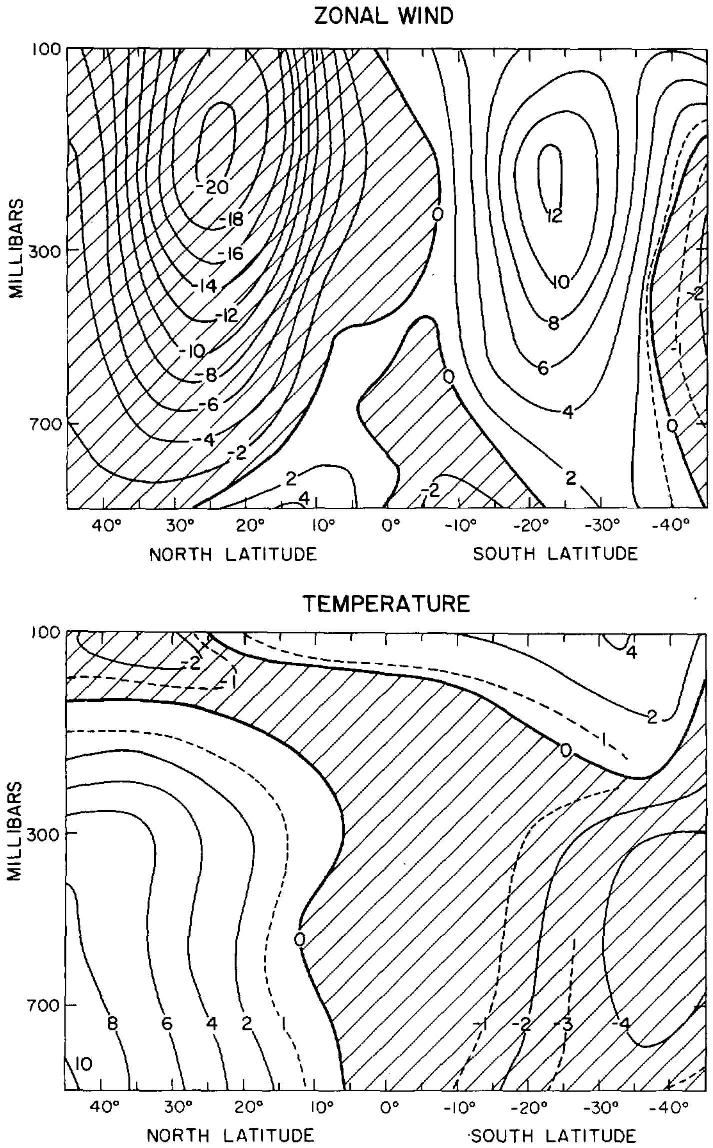


FIGURE 3.—Observed annual component (first harmonic) of the geostrophic zonal wind and temperature. Positive amplitude indicates maximum values attained in August; negative amplitudes indicate maximum values in February; units, as in figures 1 and 2.

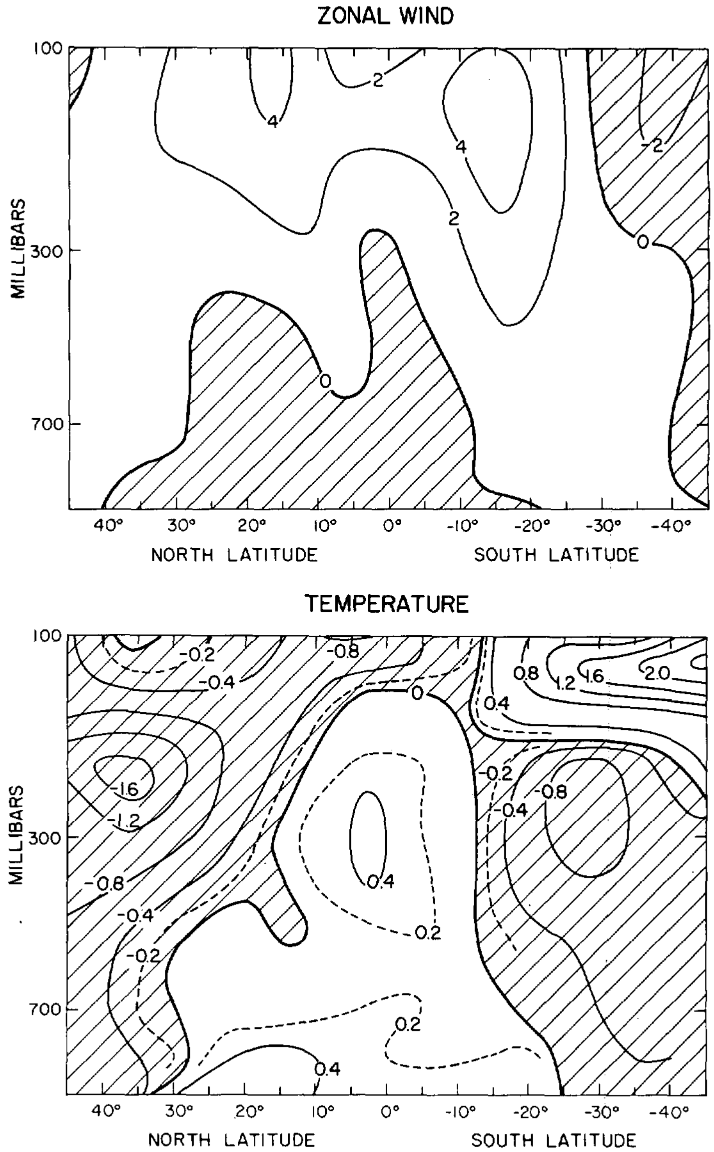


FIGURE 4.—Observed semiannual component (second harmonic) of the geostrophic zonal wind and temperature. Positive amplitude indicates maximum values attained in May and November; negative amplitudes indicate maximum values in February and August; units, as in figures 1 and 2.

either 0° or 180°, depending on whether they were originally less than 90° or not. The components with 0° phase, that is, maximum amplitude during the Northern Hemisphere winter, are then drawn with negative sign. Thus, these figures give the observed amplitudes of the harmonics but only crudely indicate the observed phases. The phases of the annual component of observed zonal wind and temperature are such that the maximum amplitudes occur predominantly during the months of August and February. Likewise, we have found in the calculations to be described here that the phase of the computed annual component of the zonal winds and temperatures lags the phase of the sources and mean meridional circulation by approximately a month. Our discussion is simplified by using "January" and "July," the times when the assumed sources have extreme values, to be synonymous with Northern Hemisphere winter and summer in describing all variables. It should be remembered that annual components of the observed and computed winds and temperatures actually have extreme values during February and August. The first harmonic largely accentuates the seasonal differences already mentioned. Note that the minimum temperatures during January which occur in the Southern Hemisphere lower stratosphere extend across the Equator well into the Northern Hemisphere (Reed and Vlcek 1969; van Loon and Jenne 1969, 1970*b*).

The positive values for the semiannual component in figure 4 indicate where the zonal winds and temperatures

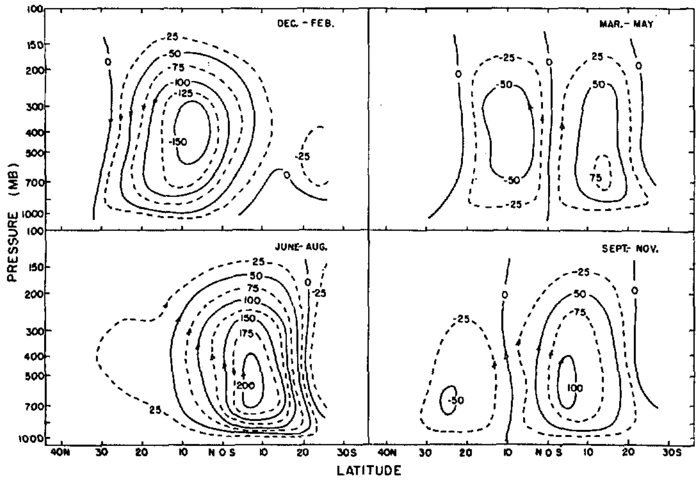


FIGURE 5.—Observed tropical mean meridional circulation with mass flux of 25×10^{12} g/s between streamlines.

at the equinoxes exceed the annual mean values, and negative values indicate where the mean of summer-winter values exceeds the annual mean. The phase of the observed second harmonic in zonal wind and temperature is such that maximum amplitudes are actually reached in the months of May and November lagging the equinoctial dates by over a month. We have also found this lag in the theoretical calculations for sources having maximum amplitudes at the equinoxes. Referring to this May–November phase, we see maximum westerlies in the upper troposphere of amplitude 3 to 4 m/s and centered about 15° from the Equator; maximum easterlies in the lower troposphere below the westerlies with amplitudes less than 1 m/s; a positive temperature in the equatorial troposphere with maximum amplitude at the Equator of 0.5°C; above this positive temperature region a negative temperature region in the lower stratosphere of maximum amplitude of about 1°C; and a negative temperature region in the subtropical troposphere of both hemispheres. On the poleward side of the subtropical negative temperatures are easterlies, the amplitude of which is about 2 m/s in the Southern Hemisphere and weaker in the Northern Hemisphere. See van Loon and Jenne (1969, 1970*a*) for further discussion of the observed semiannual oscillation in the Tropics.

Figure 5 from Newell et al. (1969) indicates the tropical mean meridional circulation as a function of season. The main features to note are the strong Hadley cells in the winter hemispheres. Their rising branches extend well into the summer hemisphere with consequent large fluxes of mass between hemispheres. The July cell is significantly stronger, and its upward branch extends further into the summer hemisphere. It is also evident that the Southern Hemisphere cell, either in the annual mean or at the equinoxes, will be the stronger of the two Hadley cells. Oort and Rasmusson (1970) have recently discussed in greater detail the seasonal variation of the mean meridional circulation.

3. DESCRIPTION OF THE MODEL

The model used here is the same as we derived in I, except for the following modifications:

1. At and above the 100-mb level, the static stability has a value of 0.025, which is five times greater than the value used for lower levels.
2. We have increased the frictional and radiative damping in the lower troposphere in a manner to be described below.
3. We have extended our source parameterization to include variation with seasons and between hemispheres, as described in the next section.

The model consists of a linearized zonal momentum equation, geostrophic wind equation, equation of hydrostatic balance, linearized thermodynamic equation, and continuity equation [cf. eq (1) through (5) of I]. These are reduced to a partial differential equation for the stream function ψ; thus

$$\frac{\partial^2\psi}{\partial z^2}+\frac{\partial\psi}{\partial z}+\left(\frac{d+i\nu}{\alpha+i\nu}\right)\frac{S}{y^2}\frac{\partial^2\psi}{\partial y^2}=\frac{e^{-z}}{2\Omega r_e y^2}\left[y\frac{\partial^2\overline{u'v'}}{\partial y\partial z}-\frac{R/c_p}{2\Omega}\left(\frac{d+i\nu}{\alpha+i\nu}\right)\frac{\partial\overline{Q}}{\partial y}\right] \tag{1}$$

corresponding to eq (6) of I. Motions are periodic in time with period ν, y=latitude, and z=log pressure^{-1}, $\overline{u'v'}$ is the horizontal eddy momentum transport, $\overline{Q}$ is the given heating, d is the Rayleigh friction, and α is the Newtonian cooling coefficient. The static stability S in eq (1) can vary with z. For simplicity, we assumed S=0.005 in I; here, we again take S to be 0.005 for the troposphere but take S to be 0.025 for the stratosphere, which is assumed to lie above 100 mb.

To obtain subtropical easterlies in which the amplitude is not grossly exaggerated, it is necessary (as discussed in I) for the lower layers of the troposphere to lose more momentum for a given wind speed than do the upper layers. We include this variation of drag with height in our model in the following crude manner. Motions are determined using eq (1), but d and α have the following variation with z:

$$\left.\begin{matrix} d=d_o \\ \alpha=\alpha_o \end{matrix}\right\}\div[0.55+0.5\tanh(z-1.5)] \tag{2}$$

where d_o and α_o are the values (19.4 day)$^{-1}$ used in I as the damping rates. Equation (2) gives a rate of damping that is a factor of 10 larger than d_o and α_o near the ground and decreases gradually to around the values of d_o and α_o at the tropopause.

Since d and α are taken to be equal, they cancel out of eq (1), so the only change from the motions that would be obtained using $d=d_o$, $\alpha=\alpha_o$ is that the zonal winds and temperatures are multiplied by the term in brackets in eq (2). The meridional circulation is unchanged. Actually, eq (1) is no longer strictly valid since its derivation requires assuming d constant. There is some justification based on scale analysis for omitting the term proportional to $d'(z)$ which is obtained in a more exact derivation, and it is easy to return to the dynamically consistent solution obtained for constant d and α by again dividing out the variable part of eq (2) from the zonal winds and temperatures. Since the actual rate of radiative damping probably does not become as large as does the frictional damping near the surface, it appears that use of the differential equation derived with variable d rather than eq (1) is not warranted without further improvements in the parameterization of the vertical variation of d and α.

Solution of eq (1) again proceeds by separation of variables and by making the appropriate eigenfunction expansions as described in I. All the formalism of the previous paper up to and including eq (20) of I is still applicable. The method of solution for the vertical structure equation is essentially the same, but the algebra is now complicated considerably by the matching to the region of higher static stability in the stratosphere. The solution of the vertical structure equation in the troposphere consists of expressions (25) through (29) of I, with the addition of further solutions to the homogeneous equation. The solution for the stratosphere must decay to zero for large z.

The coefficients of the homogeneous solutions in the troposphere and stratosphere are uniquely determined by the requirement that ψ and ψ' must be continuous across the interface. In other words, there can be no discontinuity in the vertical and meridional velocities across the tropopause. There will be, however, discontinuities of the temperature perturbation where the static stability is discontinuous.

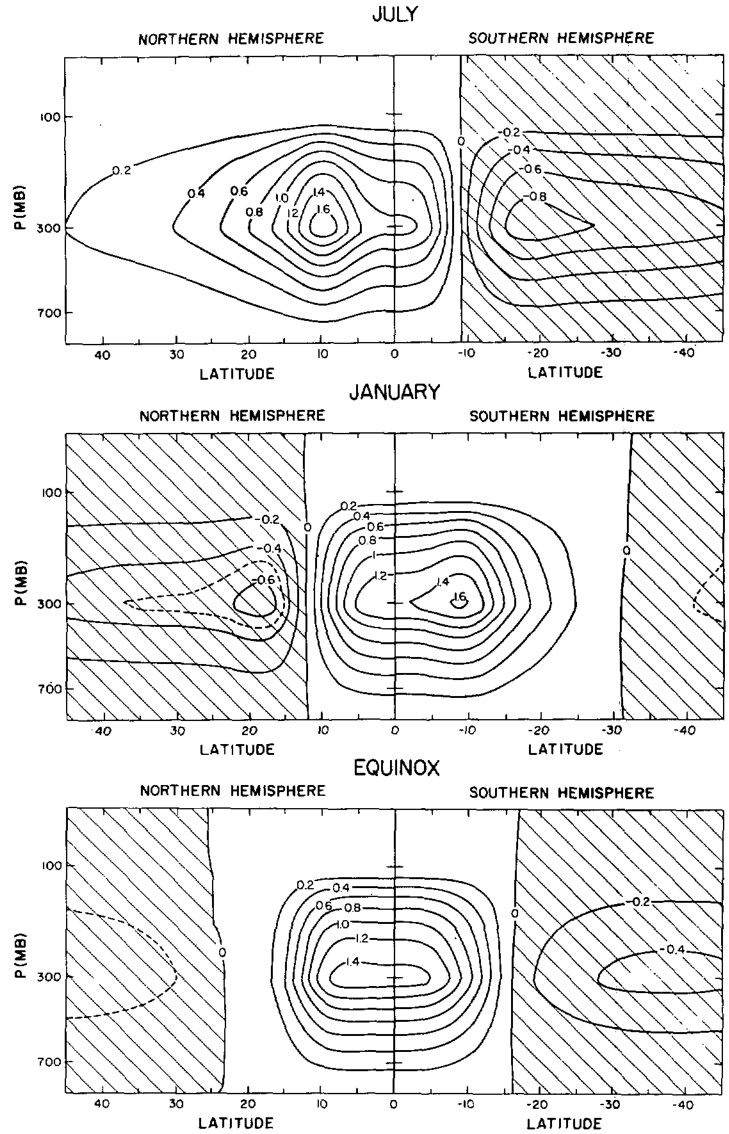


FIGURE 6.—Assumed rate of heating (in deg./day) for July, January, and the equinoctial months.

4. DESCRIPTION OF THE THERMAL SOURCE TERM

We model the latent heat release by the tropical rain belt essentially as described in I, except that now we attempt to include the variation of the latitudinal distribution of rainfall with season. Zonal mean, monthly rainfall data are not available, but the qualitative behavior of the tropical rain belt as a function of season is reasonably well known. In much of the Western Hemisphere, the tropical rain belt remains essentially in the Northern Hemisphere the whole year; whereas in the Eastern Hemisphere, where rainfall amounts are largest, a large migration between summer hemispheres is observed (Palmén and Newton 1969, p. 427). This picture is often complicated by the simultaneous appearance of rain belts on both sides of the Equator and by regions where the rainfall intensity varies with seasons but the location of maximum rainfall remains fixed.

The latitudinal distribution of the zonal mean rainfall in the Tropics according to the above description shows a rather strong maximum centered roughly in the 10° to 15°N belt in July, a quite flat maximum somewhat north of the Equator at the equinoxes, and a maximum of intermediate strength between 5°S and 10°S during January. The movement of the maximum in the zonal mean rainfall into the Southern Hemisphere during the Northern Hemisphere winter season is dictated by the large excursion into the Southern Hemisphere of the region of largest rainfall amounts in the Asiatic sector.

Figure 6 shows the distribution of zonal mean heating rates we assume to be consistent with the above picture. The two-dimensional source distribution was constructed in the manner described in I except that a more elaborate spectral weighting function was used. Since this heating is the deviation from the heating distribution that maintains a basic reference temperature, it becomes negative outside the region of large tropical rainfall. For reference, the assumed annual mean integrated amount of heat input on the Equator is 52.5 kilolangleys/yr (a kilolangley of heating corresponds to 1.67 cm of rainfall). At the equinoxes, we use a value larger by 10 percent than this value; and during January, we use a value larger by 3 percent. During July when the maximum mean rainfall is well north of the Equator, we take the rainfall at the Equator to be less by 23 percent than the annual mean. In other words, the amplitudes of the first and second harmonics of heating at the Equator are assumed to be 13 percent and 10 percent, respectively, of the annual mean values. Although these values do not seem unreasonable, it should be noted that the unavailable actual values must depend on such questions as the length of time that the maximum rainfall remains near the Equator in migrating between summer hemispheres.

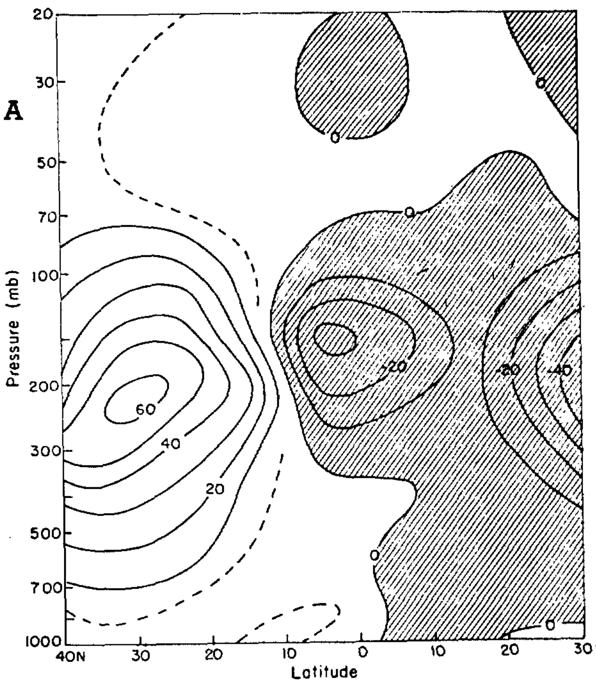
The amplitude of the actual second harmonic in rainfall is poorly known but must be much less in the zonal mean than it is locally at longitudes where the tropical rain belt does migrate smoothly between hemispheres. Chiusano (1970), who used a preliminary version of our model to study the semiannual oscillation, assumed rainfall over Africa was representative of the zonal mean. He obtained from African rainfall data a semiannual component of heating more than half as large as that used here for an annual mean value and correspondingly large wind and temperature amplitudes.

For simplicity, we take the amplitude of the semiannual component of heating to be symmetric across the Equator. Its phase is such as to give maximum heating rates during the equinoctial months. The annual wave in the heating and the annual mean heating have components asymmetric about the Equator, reflecting the difference between summer and winter hemispheres and the displacement of the meteorological Equator into the Northern Hemisphere. The asymmetric component of the annual mean heating rate is responsible for the maximum heating rates being displaced into the Northern Hemisphere during the equinoxes and for the maximum in the summer heating rate being much more pronounced in the Northern Hemisphere.

5. DESCRIPTION OF THE MOMENTUM FLUX SOURCE TERM

The momentum source term used in our model calculations is based largely on the momentum transport data described by Kidson et al. (1969). Figure 7 (from their study) shows the momentum transport which they obtained. Included in this figure is the momentum transport by the mean meridional cell between 33°N and 15°S. Since mean circulation transport is omitted in our model on the left-hand side of the zonal equation of motion, it is appropriately included with the eddy transports as a forcing term on the right-hand side. Thus, we shall regard figure 7 as the proper description of observed momentum transports in prescribing forcing for the model.

A



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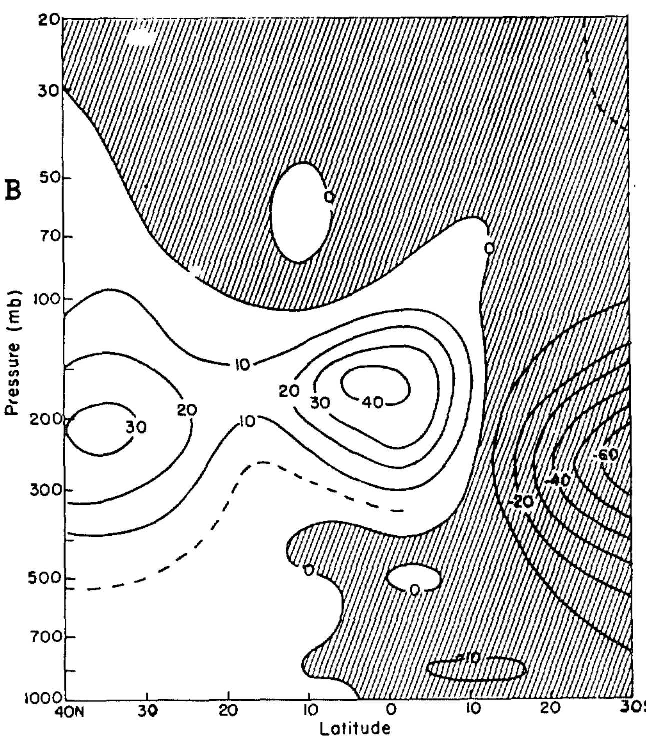

FIGURE 7.—Observed eddy momentum flux. Three-month mean values centered on (A) January and (B) July; units, m^2/s^2. Positive values denote northward flux.

Of particular interest are the strong cross-equatorial momentum fluxes into the summer hemisphere. The maximum value of this cross-equatorial flux is roughly 50 percent greater during July than during January, a difference largely due to the contribution to the flux by the equatorward mean meridional circulation in trans-

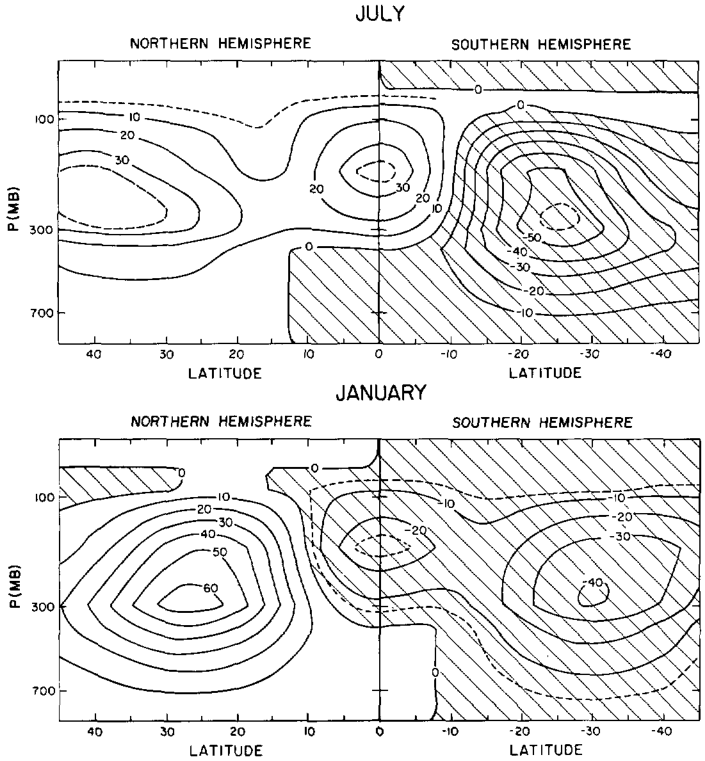

FIGURE 8.—Assumed eddy momentum transport for July and January; units, m^2/s^2. Positive values denote northward flux.

porting the momentum of the strong easterly jet of the July equatorial upper troposphere.

The mid-latitude maximum poleward fluxes are roughly twice as large in the winter hemispheres as in the summer hemispheres and occur in a region about 10° of latitude closer to the Equator in the winter than in the summer. The maximum poleward flux is significantly less in the Northern Hemisphere summer than in the Southern Hemisphere summer, whereas both winter hemispheres have essentially the same values of maximum flux. Finally, the level at which maximum transport takes place appears to decrease in height going away from the Equator.

Figure 8 shows the July and January momentum flux distributions assumed for the model calculations. We have attempted to reproduce qualitatively the features mentioned above.

The momentum source was constructed from the power series using eq (19) of I for the even component of momentum convergence and a similar expression for the odd component. Momentum sources centered at $z=1.8$ and $z=1.2$ were assumed with different coefficients of the weighting factors to reproduce the downward tilt with increasing latitude of the level of maximum momentum transport. The amplitudes of the spectral weighting factors consist of annual mean and first harmonic components.

Momentum divergence occurs in a region of the winter hemisphere extending from the Equator to about 25° with largest divergence in the subtropics. This winter hemisphere divergence is larger in the Southern Hemisphere than in the Northern Hemisphere and provides the greater cross-equatorial fluxes that occur during July. In the summer Tropics, there is a region of convergence which is larger in July than in January. In the summer subtropics, there is a region of momentum divergence which is weak relative to the divergence in the subtropics of the winter hemisphere and is weakest in the Northern Hemisphere. Finally, the regions of momentum convergence in middle latitudes are strongest in the winter hemispheres, of intermediate strength in the Southern Hemisphere summer, and weakest in the Northern Hemisphere summer.

We have used the mean of the assumed July and January momentum fluxes for calculation of the equinox zonal circulation. However, examination of some features of the semiannual oscillation of winds and temperatures in the subtropics suggests the presence of a small semiannual oscillation in the eddy momentum transport as discussed in the next section.

6. COMPUTED ZONAL MEAN STATE

RESPONSE TO HEAT SOURCE

We describe here the circulation that is forced by the heat source shown in figure 6. The results for July and January are seen in figures 9 through 12. The meridional circulation, given in figure 9 in terms of mass flux streamlines, is dominated by a large winter-hemisphere Hadley cell with much of its upward branch in the summer hemisphere. This cell is controlled by the release of latent heat in the tropical rain belt, which has been displaced into the summer hemisphere. There also occurs a weaker Hadley circulation in the summer hemisphere. In January, as a result of the less pronounced shift of the belt of maximum heating from the Equator, the summer cell transports more mass and the winter cell less mass than in July.

Figure 10 shows the northward velocity component, again illustrating the strong winter and weak summer Hadley cells. The Coriolis torque term forcing the zonal winds is proportional to the product of the velocities shown here with latitude. The computed zonal winds are shown in figure 11. The upper branch of the major Hadley cell forces an easterly jet near the Equator in the summer Tropics and a strong westerly jet in the winter Tropics. The lower return branch of the major Hadley cell drives equatorial westerlies below the easterly jet and subtropical easterlies below the westerly jet. All these wind systems just mentioned are stronger in July than in January because of the stronger winter meridional cell. In the subtropics of the summer hemisphere, the weak meridional cell gives westerlies aloft and easterlies in the lower troposphere. These winds are weakest in July.

The temperature perturbations we obtain are shown in figure 12. The region of maximum temperature in the summer troposphere extends beyond the latitude of

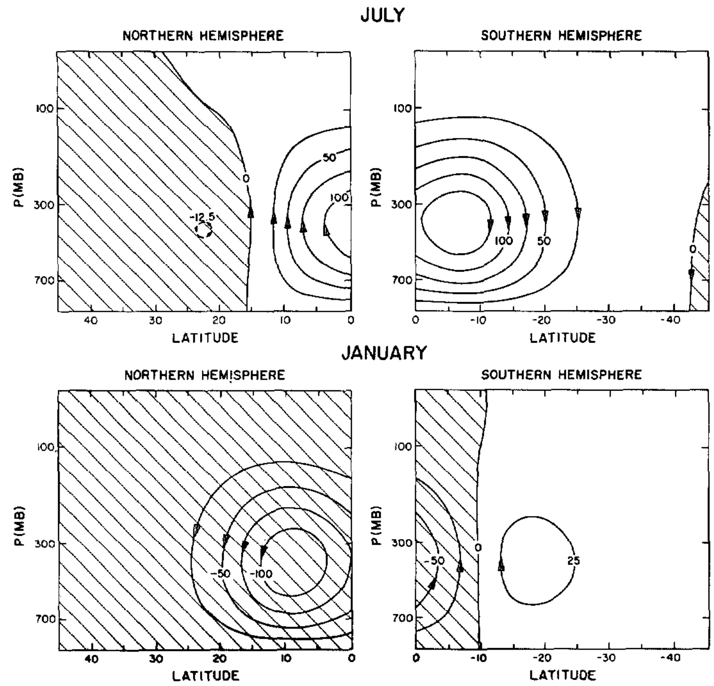

FIGURE 9.—Streamlines of meridional circulation forced by heat source, July and January; units, 10^{12} g/s.

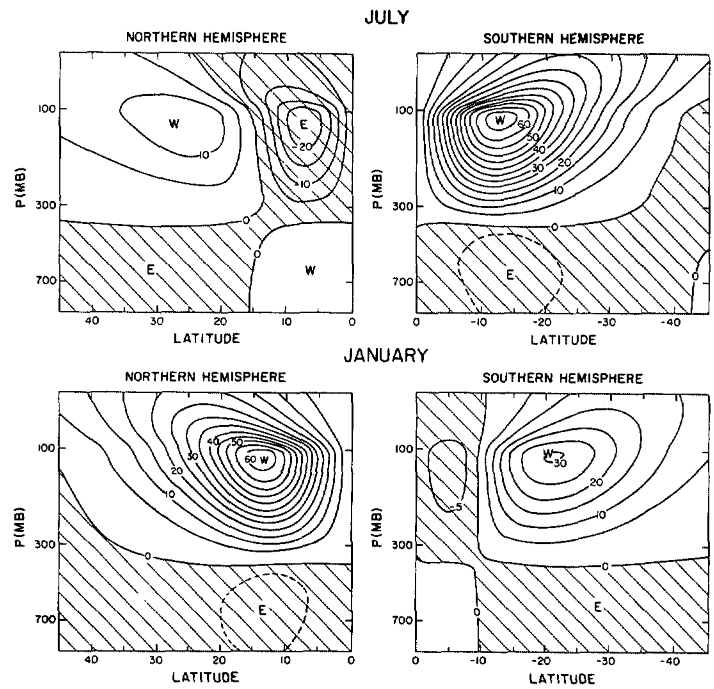

FIGURE 11.—Zonal wind forced by heat source, July and January; units, m/s.

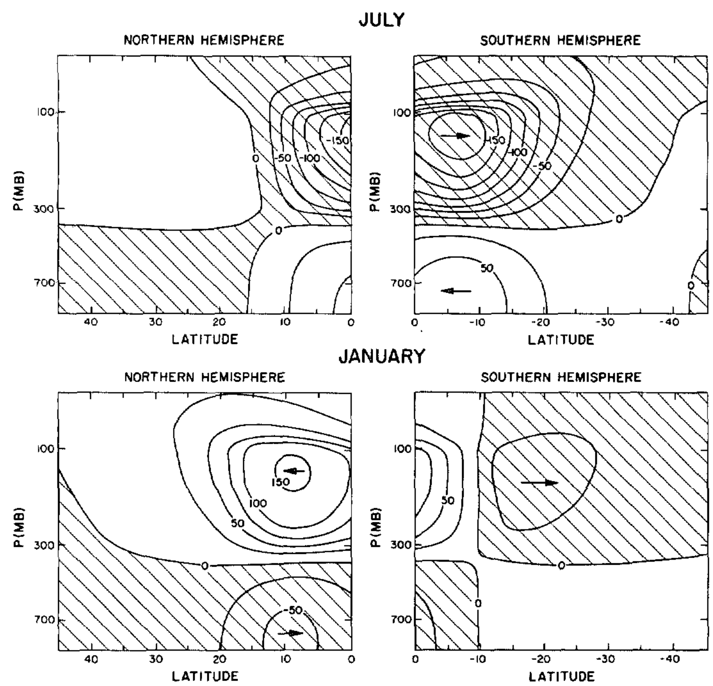

FIGURE 10.—Meridional wind forced by heat source, July and January; units, cm/s.

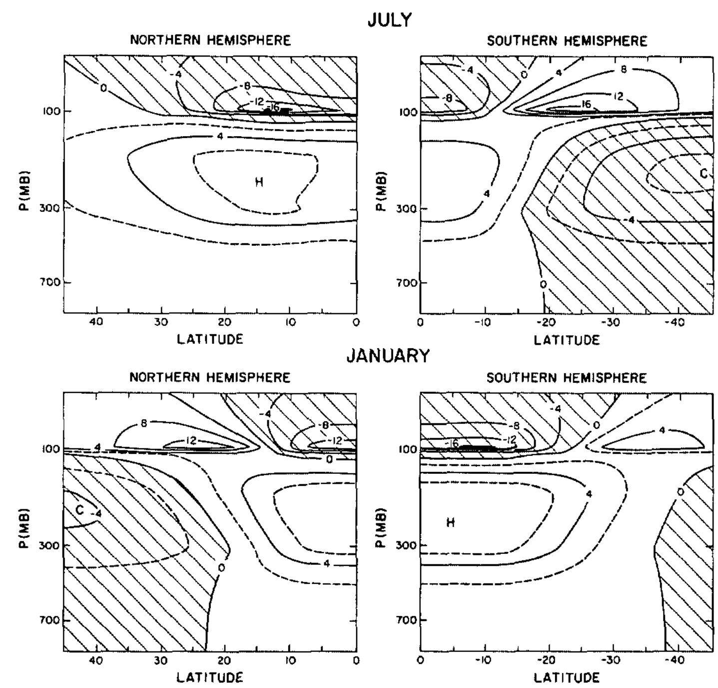

FIGURE 12.—Temperature perturbation forced by heat source, July and January; units, °C.

maximum heating because vertical motions decrease with latitude on the source's poleward side more rapidly than does the heat source itself. The welling up from the source region into the region of high static stability gives large adiabatic cooling and minimum temperatures above the source in the lower stratosphere. Similarly, the downward branch of the major meridional cell creates a region of maximum warmth in the lower stratosphere in the subtropics which is warmer in July than in January because of the greater strength of the July meridional cell. The greater displacement of the upward branch of the July cell into the summer hemisphere results in less vertical motion

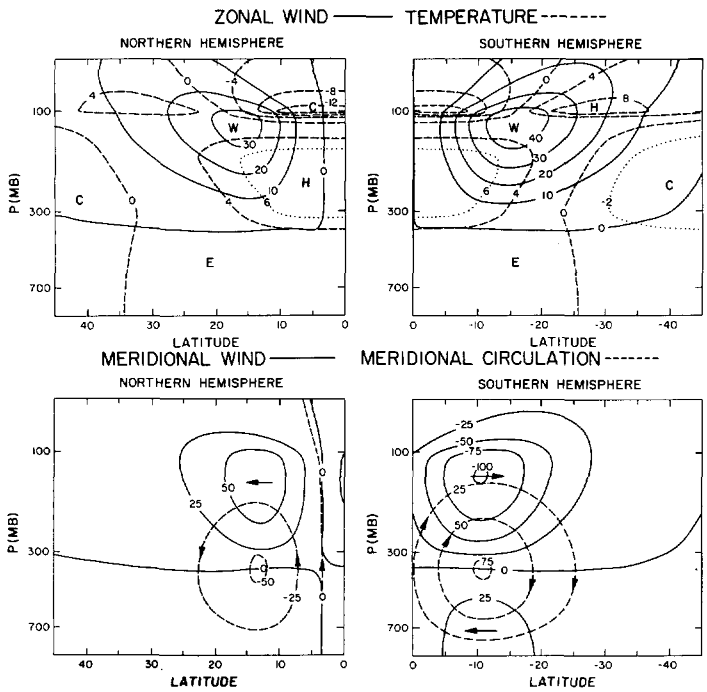

FIGURE 13.—Annual mean response to thermal source. The top frames show zonal wind in solid lines and temperatures in dashed lines. The bottom frames show meridional wind in solid lines and streamlines of the meridional circulation in dashed lines. The units are the same as those in previous figures.

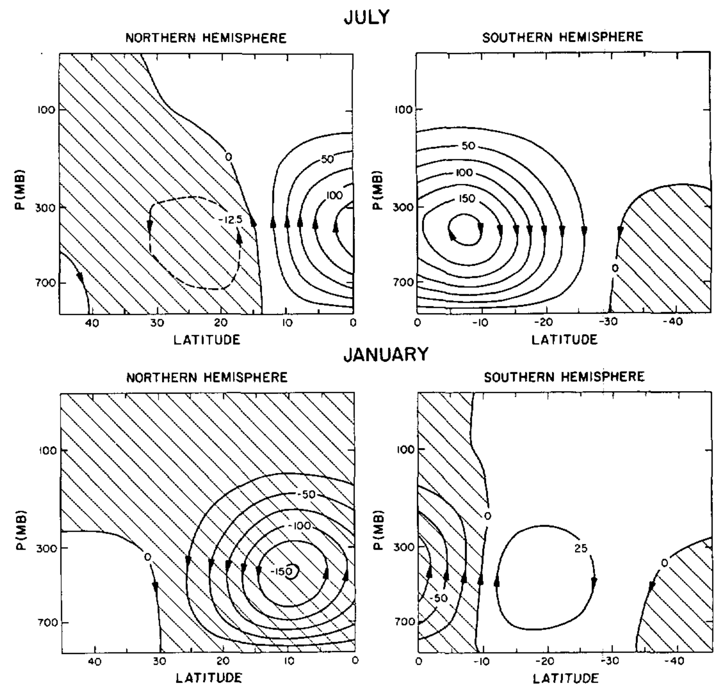

FIGURE 14.—Streamlines of meridional circulation forced by heat and momentum sources, July and January; units, 10^{12} g/s.

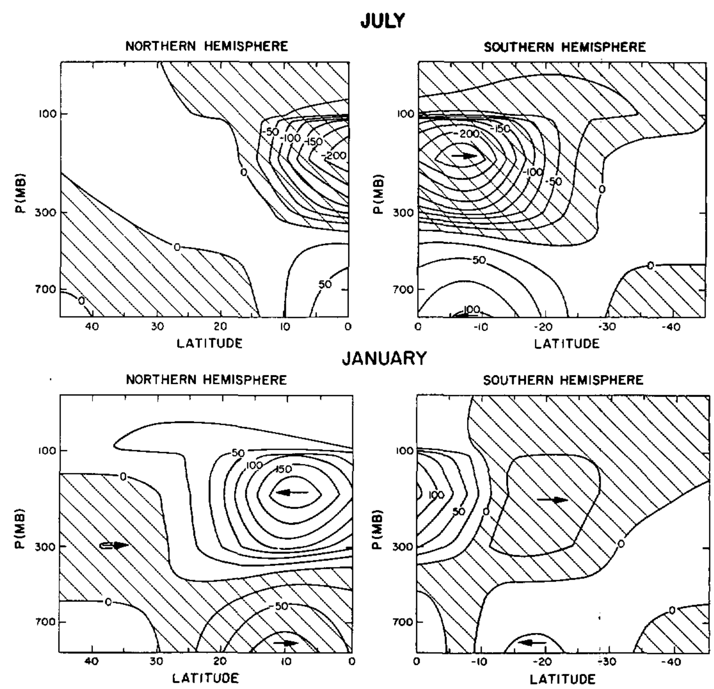

FIGURE 15.—Meridional wind forced by heat and momentum sources, July and January; units, cm/s.

at the Equator and hence in a warmer equatorial lower stratosphere in July than in January, as suggested by van Loon and Jenne (1970*b*).

The annual mean thermal source for the model consists of the mean of July, January, and two of the equinox source distributions shown in figure 6. The response to an annual mean heating is summarized in figure 13 and reflects the location of the "meteorological Equator" northward of the geographical Equator. The Southern Hemisphere Hadley cell is considerably stronger than the Northern Hemisphere cell. This difference is accompanied by the indicated variation between hemispheres of the mean zonal winds and temperatures.

RESPONSE TO HEAT AND MOMENTUM SOURCES

The modifications which result from the addition to the July–January solutions just described of the response to the eddy momentum sources are shown in figures 14 through 17. The major meridional cells, as shown in figures 14 and 15, are of approximately one-third greater strength than those driven by the heat sources alone. Ferrel cells now occur in middle latitudes. They remove momentum from the upper troposphere where it is brought by the eddy transports and add momentum to the lower troposphere to maintain the westerlies there against friction.

The large momentum divergence from the subtropics of the winter hemisphere greatly enhances the amplitude of the subtropical easterlies as seen by comparing figure 16 with figure 11. The winter westerlies are shifted to a latitude 30° to 35° from the Equator. The weaker summer westerlies are likewise shifted poleward to a latitude of about 45°.

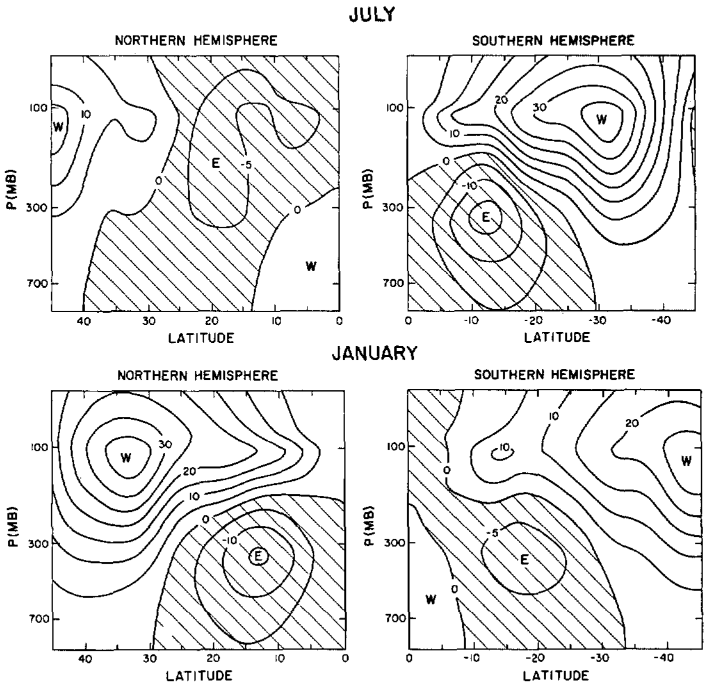

FIGURE 16.—Zonal wind forced by heat and momentum sources, July and January; units, m/s.

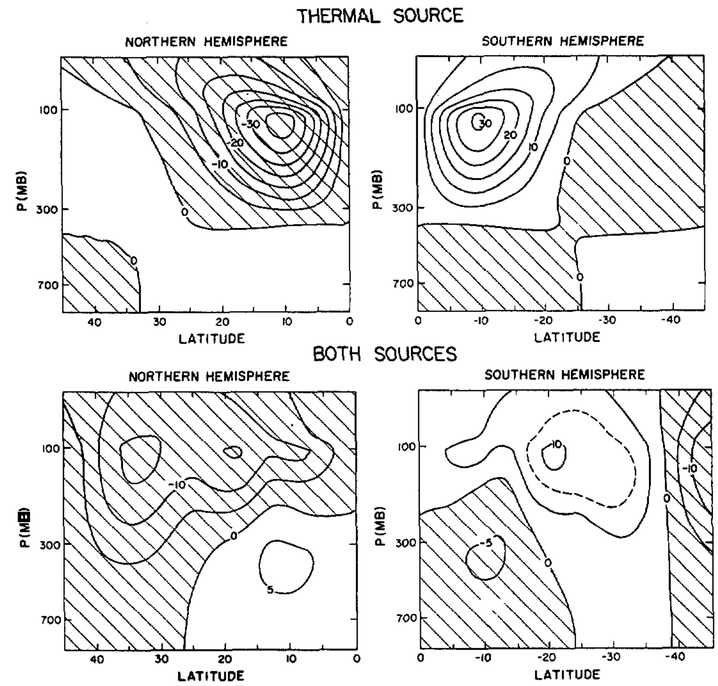

FIGURE 18.—Annual component (first harmonic) of the zonal wind. The top frame shows the response to heat source and the bottom frame shows the response to both heat and momentum sources; units, m/s.

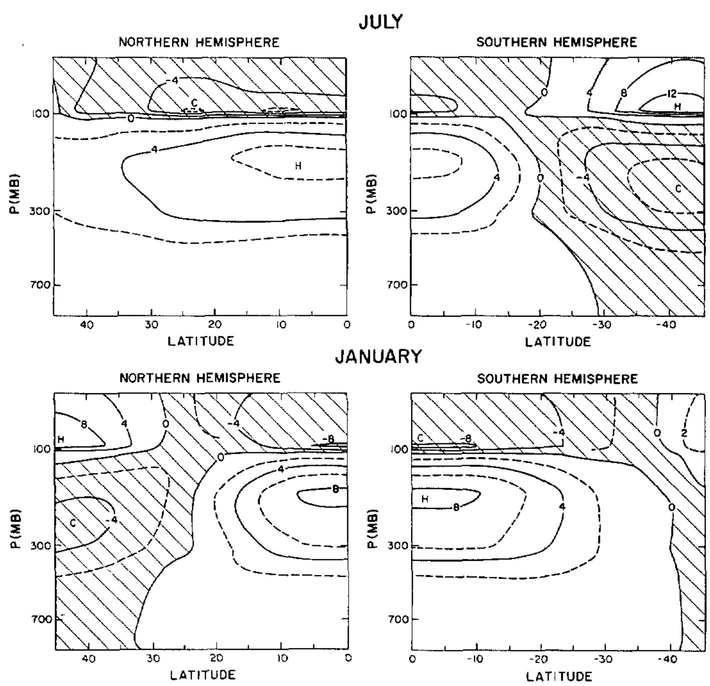

FIGURE 17.—Temperature perturbation forced by heat and momentum sources, July and January; units, °C.

As the horizontal branch of the meridional circulation forced by momentum transports redistributes momentum by Coriolis torques in the vertical, so also the vertical branch redistributes heat horizontally by adiabatic warming. We see, in particular, in comparing figures 12 and 17 that the vertical motions have shifted the region of greatest warmth in the lower stratosphere of the winter hemisphere from the subtropics to middle latitudes. In the winter troposphere, the zone of maximum baroclinicity also shifts poleward as it must with the poleward shift of the westerlies resulting from the eddy momentum fluxes. The momentum fluxes force downward motion in the equatorial lower stratosphere which partially cancels the upward motion forced by the tropical rain belt. This downward motion is greatest in the summer hemisphere near the latitude where the poleward momentum flux is minimum.

The annual components of wind and temperature forced by the heating alone and by both sources are compared in figures 18 and 19. The greater amplitude of the first harmonic of the zonal westerlies in the Northern Hemisphere compared to the Southern Hemisphere reflects the relatively weak summer westerlies occurring in July. We need not comment on previously mentioned features, but it is appropriate here to discuss this difference between the summer zonal winds in middle latitudes of the two hemispheres. The weak July summer westerlies as obtained here are a consequence of a relatively small momentum convergence in middle latitudes of the Northern Hemisphere at that time. This weak convergence appears to be consistent with the relatively strong winter cell and relatively weak summer cell. That is, the westerly momentum made available to the summer upper troposphere by the meridional circulation is minimum during July.

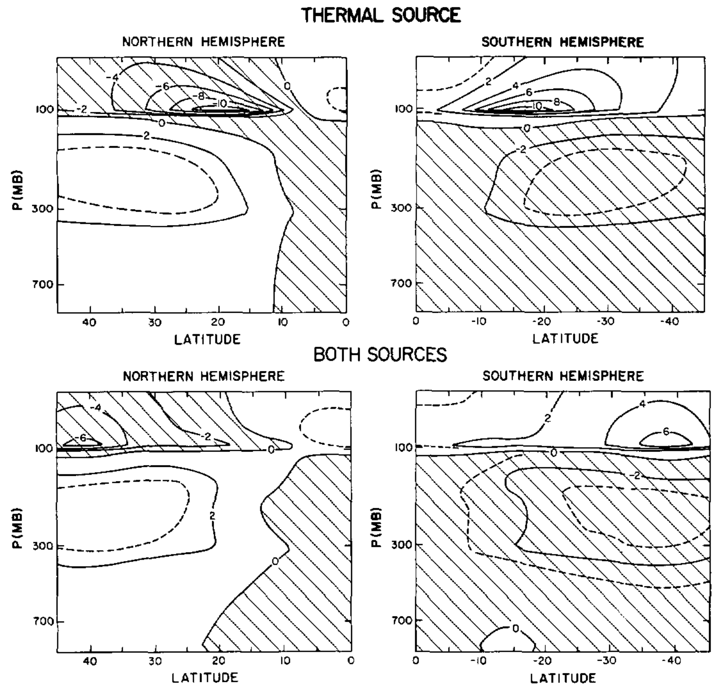

FIGURE 19.—Annual component (first harmonic) of the temperature (arranged the same as fig. 18); units, °C.

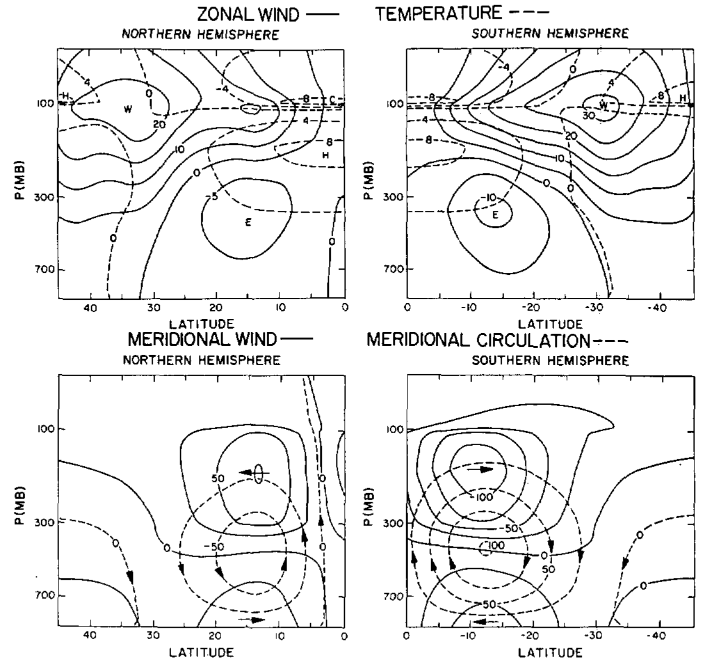

FIGURE 20.—Equinox response to heat and momentum sources (arranged the same as fig. 13).

The lower temperatures during January than during July in the lower stratosphere extend farther into the Northern Hemisphere when the momentum-forced circulation is included. This effect depends on the momentum convergence in the summer Tropics which accompanies the large momentum flux into the summer hemisphere. Our calculated annual temperature wave in the mid-latitude lower stratosphere does not have an amplitude significantly less in the Northern Hemisphere than in the Southern Hemisphere such as is observed. Adiabatic cooling in the stratosphere resulting from the upward branch of the major Hadley cell extends out to middle latitudes in July according to the model calculations. This cooling and hence also the coldness of the July stratosphere away from the Tropics in the Northern Hemisphere is unrealistic. The model for this reason greatly overestimates the amplitudes of the annual component of temperature in the Northern Hemisphere lower stratosphere. This defect is connected with the inability of the model driven only by tropospheric sources to generate in July tropical easterlies which have amplitudes increasing upward into the stratosphere as occurs in the zonal mean. The easterly jet core with maximum winds at 100–150 mb is similar to the tropical easterly jet stream observed over the Asiatic sector (Palmén and Newton 1969). Temperature amplitudes at the tropopause are exaggerated, in general, by the model because of the discontinuity in stability there.

The circulation obtained for the equinoxes is shown in figure 20. Again, the Southern Hemisphere has a stronger Hadley cell in the Tropics and stronger zonal westerlies in mid-latitudes. The anomalous warmth in the mid-latitude lower stratosphere is greater in the Southern Hemisphere.

SEMIANNUAL OSCILLATION IN WINDS AND TEMPERATURES

A semiannual component in zonal mean rainfall at the Equator, with amplitude reaching its maximum near the equinoxes, results from the equatorial crossing of the mean tropical rain belt as it migrates twice a year between hemispheres. To evaluate the hypothesis that the semiannual winds and temperatures in the Tropics are a consequence of the seasonal migration of the tropical rain belt, we assume the same distribution of heating as was used to calculate the component of the annual mean winds symmetric about the Equator. The results given by our model for a heating amplitude which is 0.1 of that used to give the annual mean symmetric response are shown on the left-hand side of figure 21. Most of the observed features of the semiannual components as previously described are thus reproduced. The circulation is essentially the same except for amplitudes as the average of the two sides of figure 13. There is some indication in figure 4 that the regions of largest semiannual temperature range are shifted into the Northern Hemisphere, in which case we would expect the qualitative differences between hemispheres indicated in figure 13 to exist also in the semiannual circulation.

To obtain the observed maximum in the out-of-phase temperature in the subtropical troposphere and the out-of-phase winds poleward of this maximum, it was found necessary to hypothesize a second harmonic com-

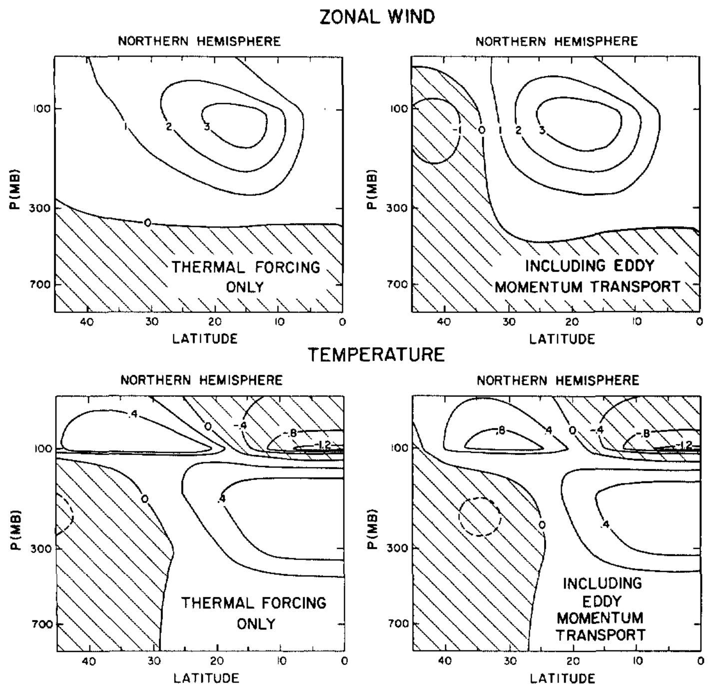

FIGURE 21.—Semiannual component (second harmonic) response. The top frames show zonal wind (m/s) and the bottom frames show temperature (°C). The left-side frames show response to only thermal forcing and the right-side frames show response when eddy momentum transport is included.

ponent in the eddy momentum fluxes. If we assume the amplitude of the mean poleward transport at the solstices averaged between hemispheres exceeds that at the equinoxes, we get relative momentum convergence at the equinoxes in the Tropics and a divergence in middle latitudes. The right-hand side of figure 21 shows the results for such a calculation when the mean momentum flux at the solstices exceeds the mean poleward flux at the equinoxes by roughly 0.1.

7. DISCUSSION

Our calculations indicate how basic features of the annual variation of the zonal winds and temperatures from the Tropics to mid-latitudes can be reproduced in terms of the annual variation of tropical rainfall and of the horizontal eddy flux of momentum. Other factors such as the annual variation of direct heating by solar radiation seem less significant, at least in the Tropics. There is clearly a need for further observational and theoretical understanding of both the distribution of rainfall and the horizontal eddy momentum fluxes in the Tropics as well as their differences between hemispheres. It should be emphasized that studies of the seasonal variation of tropical rainfall must recognize the large longitudinal asymmetries in describing the zonal mean values. The longitudinal asymmetries in rainfall will drive planetary wave components of the circulation. Although these components cannot be deduced from the present theory, they must be added to the zonal mean values if the calculated circulation at a given longitude is to be compared with observation.

We have obtained the greater strength of the Southern Hemisphere summer zonal westerlies relative to those of the Northern Hemisphere as a direct consequence of greater eddy momentum transport from the Tropics to middle latitudes in the Southern Hemisphere summer. The stronger winds in the Southern Hemisphere summer are necessarily accompanied by a stronger poleward decrease of temperature below the maximum winds. This stronger temperature gradient is produced in our model in the troposphere largely by the gradient of adiabatic cooling due to the vertical branch of the meridional circulation forced by the eddy fluxes. This greater baroclinicity in the Southern Hemisphere matches the greater poleward temperature decrease at the surface. That is, the largely oceanic lower boundary in the Southern Hemisphere summer at middle latitudes imposes a larger poleward temperature gradient on the atmosphere than does the largely continental Northern Hemisphere surface in July.

A greater poleward decrease of temperature at the surface should lead to greater baroclinic development of nonzonal disturbances by baroclinic instability and, in turn, increased convergence of eddy momentum fluxes in the mid-latitude upper troposphere. The surface temperature gradient in the Southern Hemisphere summer can, through the eddy momentum fluxes, create a meridional circulation that produces the observed temperature gradient in the troposphere except near the surface where mean vertical motions are small. Direct heating of the atmosphere from the surface is probably important for maintaining the baroclinicity of the lowest layers.

The strength of the larger July Hadley cell and the weakness of the smaller July cell are due, at least in part, to the relatively large heating imposed on the atmosphere out to middle latitudes in the Northern Hemisphere as shown in figure 6. The actual heating is not very well modeled away from the Equator but should, like the assumed heating, be greater at mid-latitudes in the Northern Hemisphere in July than in the Southern Hemisphere in January because of the relatively greater continentality of the Northern Hemisphere. Since the mean displacement of the tropical rain belt into the Northern Hemisphere also can be ascribed, at least in part, to the difference in the relative amount of land in the two hemispheres, our interpretation supports the contention that the difference between the summer zonal circulations derives largely from the difference in continentality of the two hemispheres. However, the detailed mechanisms giving the differences between hemispheres according to our model involve a chain of dynamic processes not recognized in classic climatological arguments.

8. CONCLUSIONS

In passing from summer to winter, the divergence of horizontal eddy momentum flux in the Tropics increases as does the convergence in mid-latitudes. These increases accompany a large increase in the Hadley circulation bringing momentum into the tropical upper troposphere and an increase in the strength of the zonal westerlies. The equatorward shift in winter of the latitude of maximum momentum flux is accompanied by an equatorward shift of the maximum momentum convergence and produces in the model the observed shift of the zonal westerlies in the same direction. The zonal westerlies of the Northern Hemisphere summer are weak compared to those of the Southern Hemisphere summer, according to our model calculations, because of the small amount of momentum convergence in middle latitudes in conjunction with the small amount of westerly momentum made available in the summer Tropics by the meridional circulation.

Minimum temperatures in the equatorial stratosphere during January and maximum temperatures in middle latitudes in the winter stratosphere, which are more pronounced in the Southern Hemisphere during July, accompany a winter Hadley cell in July which transports more mass, and has an upward branch which is displaced further into the summer hemisphere, while not extending as high into the stratosphere as does that of the January cell. Differences between hemispheres in the Tropics in both latent heat release and momentum fluxes are required to produce these Hadley cell differences. The semiannual component of winds and temperatures in the Tropics can be largely reproduced by a second harmonic in rainfall centered at the Equator with a zonally averaged amplitude of about 10 percent of the annual mean amplitude. The semiannual wind and temperature components in the subtropics and middle latitudes, especially in the Southern Hemisphere, appear to require a semiannual eddy momentum flux for their explanation.

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